

# Detrital zircon evidence for non-Laurentian provenance, Mesoproterozoic (ca. 1490–1450 Ma) deposition and orogenesis in a reconstructed orogenic belt, northern New Mexico, USA: Defining the Picuris orogeny

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## ABSTRACT

Detrital zircon and igneous zircon U-Pb ages are reported from Proterozoic metamorphic rocks in northern New Mexico. These data give new insight into the provenance and depositional age of a >3-km-thick metasedimentary succession and help resolve the timing of orogenesis within an area of overlapping accretionary orogens and thermal events related to the Proterozoic tectonic evolution of southwest Laurentia. Three samples from the Paleoproterozoic Vadito Group yield narrow, unimodal detrital zircon age spectra with peak ages near 1710 Ma. Igneous rocks that intrude the Vadito Group include the Cerro Alto metadacite, the Picuris Pueblo granite, and the Peñasco quartz monzonite and yield crystallization ages of  $1710 \pm 10$  Ma,  $1699 \pm 3$  Ma, and  $1450 \pm 10$  Ma, respectively. Within the overlying Hondo Group, a metamorphosed tuff layer from the Pilar Formation yields an age of  $1488 \pm 6$  Ma and represents the first direct depositional age constraint on any part of the Proterozoic metasedimentary succession in northern New Mexico. Detrital zircon from the overlying Piedra Lumbre Formation yield a minimum age peak of 1475 Ma, and ~60 grains (~25%) yield ages between 1500 Ma and 1600 Ma, possibly representing non-Laurentian detritus originating from Australia and/or Antarctica. Detrital zircons from the basal metaconglomerate and the middle quartzite member of the Marqueñas Formation yield minimum age

peaks of 1472 Ma and 1471 Ma, consistent with earlier results. We interpret the onset of ca. 1490–1450 Ma deposition followed by tectonic burial, regional  $Al_2SiO_5$  triple-point metamorphism, and ductile deformation at depths of 12–18 km to reflect a Mesoproterozoic contractional orogenic event, possibly related to the final suturing of the Mazatzal crustal province to the southern margin of Laurentia. We propose to call this event the Picuris orogeny.

## INTRODUCTION

Proterozoic orogenic belts in southern Laurentia reflect a long-lived convergent margin that was active between 1800 and 1000 Ma (Karlstrom et al., 2001; Whitmeyer and Karlstrom, 2007). These orogenic belts likely extended to Baltica and other formerly adjacent continents such as Australia or Antarctica. However, the configuration of cratons adjacent to western Laurentia during the Proterozoic is widely debated (Dalziel, 1991; Karlstrom et al., 2001; Wingate et al., 2002; Betts et al., 2008, 2011; Payne et al., 2009; Boger, 2011; Evans and Mitchell, 2011; Doe et al., 2012). The southward growth of Laurentia during this 800 m.y. period is associated with two cycles of accretionary orogenesis, including the ca. 1780–1700 Ma Yavapai and ca. 1680–1600 Ma Mazatzal orogenies (Karlstrom and Bowring, 1988; Bowring and Karlstrom, 1990; Whitmeyer and Karlstrom, 2007), which produced a >1000-km-wide zone of largely juvenile crust with some intermixed older crustal components (Fig. 1) (Hawkins et al., 1996; Hill and Bickford, 2001; Jessup et al., 2005; Bickford and Hill, 2007; Whitmeyer and Karlstrom, 2007). These Paleoproterozoic accretionary events were followed by an enigmatic trans-Laurentian magmatic event involving widespread granitic magmatism

ca. 1480–1360 Ma (Anderson, 1989; Bickford and Anderson, 1993; Windley, 1993; Anderson and Morrison, 2005; Goode et al., 2008) and culminated in the Grenville orogeny (Rivers, 1997; Mosher, 1998) and assembly of the supercontinent Rodinia at 1100–1000 Ma (Dalziel, 1991; Moores, 1991).

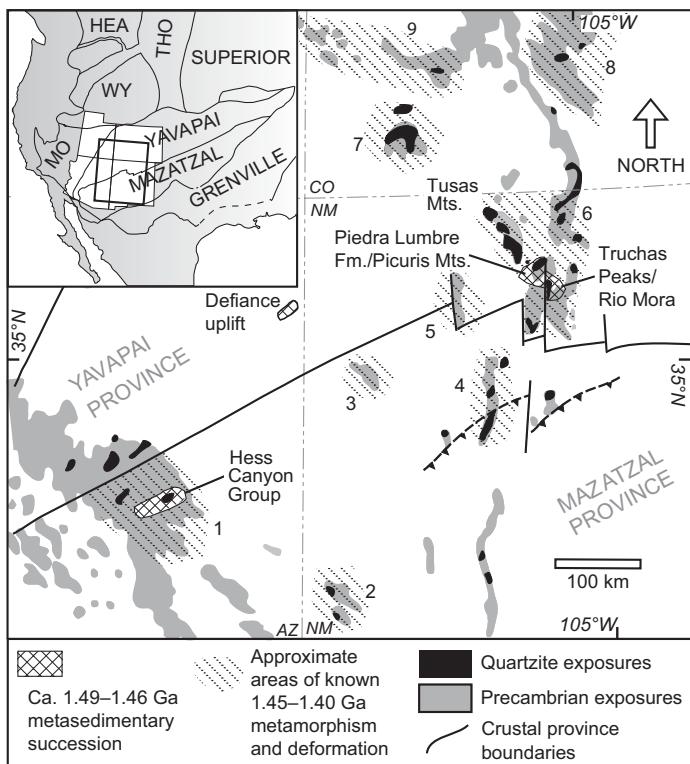
Thick metasedimentary successions exposed throughout the southwestern United States include abundant quartz arenite (Fig. 1) and locally voluminous rhyolite and serve as key marker units for distinguishing the age and tectonic setting of events that shaped the southern Laurentia margin during the Paleoproterozoic and Mesoproterozoic (Dott, 1983; Hoffman, 1988; Karlstrom and Bowring, 1988; Cox et al., 2002; Medaris et al., 2003; Jessup et al., 2006; Rainbird and Davis, 2007; Whitmeyer and Karlstrom, 2007; Amato et al., 2008, 2011; Jones et al., 2009, 2011; Doe et al., 2012, 2013). Recent work of Jones et al. (2011) and Doe et al. (2012, 2013) has established that metasedimentary successions exposed in both the Yavapai and Mazatzal Provinces that were previously interpreted as Paleoproterozoic are actually Mesoproterozoic in age, requiring significant revision of some regional lithostratigraphic correlations and tectonic models.

This study expands upon the work of Jones et al. (2011) and presents U-Pb zircon geochronology results from metasedimentary, metavolcanic, and plutonic rocks exposed in the Picuris Mountains (Fig. 1). The data reveal a previously unrecognized interval of Mesoproterozoic deposition that precedes the ca. 1450 Ma Marqueñas Formation (Jones et al., 2011). Together with new U-Pb zircon ages of interlayered and crosscutting igneous units, our findings show that (1) presumed Paleoproterozoic metasedimentary rocks of the upper Hondo Group are instead ca. 200 m.y. younger than previously interpreted; (2) Proterozoic

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**Figure 1.** Map of exposed Proterozoic rock in the southwestern United States showing crustal province boundaries, and Paleoproterozoic and Mesoproterozoic quartzite deposits. Diagonal line pattern corresponds to the general area of known ca. 1450–1400 Ma deformation and/or metamorphism in Mesoproterozoic and Paleoproterozoic rocks: 1—eastern Arizona (Doe et al., 2012, 2013); 2—Burro Mountains (Amato et al., 2011); 3—Zuni Mountains (Strickland et al., 2003); 4—Sandia and Manzano Mountains (Kirby et al., 1995; Marcoline et al., 1999); 5—Nacimiento Mountains (Kellogg and Premo, 2005); 6—Cimarron/Taos Mountains (Grambling and Dallmeyer, 1993; Pedrick et al., 1998); 7—Needle Mountains (Hunter and Andronicos, 2013); 8—Wet Mountains (J.V. Jones, Jones et al., 2010); 9—Black Canyon/Uncompahgre area (Jessup et al., 2006). Inset shows the approximate area of Figure 1 relative to the southwestern United States and the southwest United States with respect to major crustal province boundaries: MO—Mojave Province; WY—Wyoming Province; HEA—Hearne Province; THO—Trans-Hudson orogen.

rocks in northern New Mexico did not remain at midcrustal depths for ca. 200 m.y. following the Mazatzal orogeny, as proposed by previous studies; (3) a previously unrecognized Mesoproterozoic clastic deposystem across northern New Mexico was potentially connected with contemporaneous basins in Arizona (Doe et al., 2012, 2013) and with exotic, non-Laurentian sediment sources such as Australia or Antarctica (Betts et al., 2008; Boger, 2011; Doe et al., 2012, 2013); and (4) regional  $\text{Al}_2\text{SiO}_5$  triple-point metamorphism and north-vergent, fold-and-thrust-style deformation across north-

central New Mexico cannot be attributed to the Mazatzal orogeny but must postdate ca. 1457 Ma (Fig. 2; Williams, 1991; Williams et al., 1999; Karlstrom et al., 2004; Daniel and Pyle, 2006; Whitmeyer and Karlstrom, 2007; Jones et al., 2011; Cao and Fletcher, 2012).

Our new detrital zircon data require that deposition of the Pilar and Piedra Lumbre Formations occurred between ca. 1490 Ma and ca. 1450 Ma, and was followed by regional deformation, crustal thickening, plutonism, and metamorphism at midcrustal depths (Daniel and Pyle, 2006). These processes are the hall-

marks of an orogenic event that we propose to call the Picuris orogeny. We present speculative tectonic models involving a ca. 1490–1400 Ma contractional orogenic event in a collisional to transpressional tectonic setting (Nyman et al., 1994; Slagstad et al., 2009), raising the possibility that the Proterozoic growth and stabilization of southern Laurentia were more prolonged and spatially complex than previously interpreted.

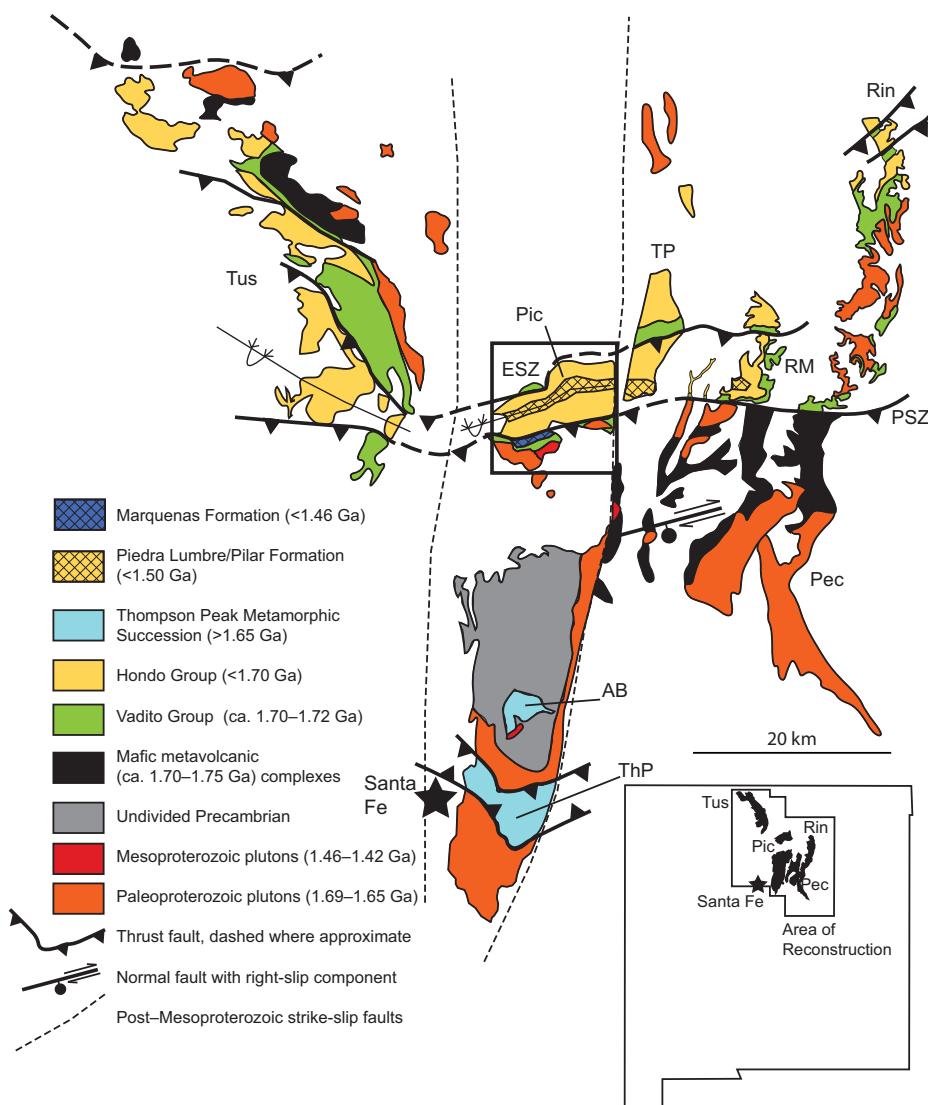
## GEOLOGIC BACKGROUND

### Proterozoic Lithostratigraphy

The oldest rocks exposed in northern New Mexico include arc-related rocks that represent the southern margin of the Yavapai crustal province and that formed between ca. 1760 and 1720 Ma (Fig. 1) (Karlstrom et al., 2004; Whitmeyer and Karlstrom, 2007). These rocks include the Gold Hill, Pecos, and Moppin mafic metavolcanic complexes, and they are interpreted to be the “basement” to ca. 1700 Ma and younger supracrustal successions such as the Vadito and Hondo Groups and the Mesoproterozoic Marqueñas Formation (Figs. 2 and 3) (Bauer and Williams, 1989). The Paleoproterozoic Vadito Group includes a complexly interlayered succession of mafic and felsic metavolcanic rocks, schist, and quartzite with minor pebble conglomerate that are crosscut by variably deformed granitic intrusions (Bauer and Williams, 1989; Bauer, 1993). This mixed volcanic and sedimentary unit was interpreted to be the result of continental rifting along either a continental margin or an intracontinental setting (Mawer et al., 1990) or an arc-related setting (Soegaard and Eriksson, 1986).

The Hondo Group overlies the Vadito Group and consists of >2.5 km of metasedimentary rock (Soegaard and Eriksson, 1985, 1986; Bauer, 1988, 1993). In the Picuris Mountains and southern Rio Mora areas, the contact between the Vadito and Hondo Groups is tectonic (Grambling and Codd, 1982; Bauer, 1993). However, in the Tusas Mountains and the northern Rio Mora area, the contact appears to be only weakly tectonized and is interpreted as a gradational, conformable succession (Bauer and Williams, 1989; Mawer et al., 1990; Williams, 1991). The lower Hondo Group (Fig. 3) is made up of 1–1.5 km of aluminous quartzite of the Ortega Formation followed by ~725 m of interbedded schist and quartzite of the Rinconada Formation (Bauer, 1988). These formations were interpreted as inner-shelf and deltaic deposits, respectively (Soegaard and Eriksson, 1985, 1989).

The upper Hondo Group consists of the Pilar and Piedra Lumbre Formations (Fig. 3). The Pilar Formation is a distinctive, fine-grained,



**Figure 2. Proterozoic fold-and-thrust belt across north-central New Mexico reconstructed to account for post-Mesoproterozoic strike-slip offset (modified from Karlstrom and Daniel, 1993; Daniel et al., 1995; Cather et al., 2006). The Pilar and Piedra Lumbre Formations exposed in the Picuris are also exposed in the Truchas Peaks and Rio Mora areas to the east. Outlined box around the Picuris Mountains shows the approximate area of Figure 3. Abbreviations: Tus—Tusas Mountains, Pic—Picuris Mountains, TP—Truchas Peaks area, RM—Rio Mora area, Pec—Pecos area, Rin—Rincon Mountains, AB—Aspen Basin area, ThP—Thompson Peak area, ESZ—Embudo shear zone, PSZ—Plomo/Pecos shear zone. Inset shows the approximate outline of the reconstructed area and the present-day locations of selected mountain ranges within the reconstructed area.**

black, carbonaceous phyllite or schist with centimeter- to meter-scale, white schistose layers (Long, 1976; Bauer, 1988) that we interpret as metamorphosed tuff beds. This formation is also exposed in the Truchas Peaks and Rio Mora areas (Grambling and Codding, 1982). The uppermost Pilar Formation grades into schist and quartzite of the overlying Piedra Lumbre Formation, characterized by centi-

meter-scale graded beds and centimeter- to meter-scale cross-bedded quartzite layers that become more abundant and thicker toward the top of the exposed section (Long, 1976; Bauer, 1988). The uppermost exposed rocks consist of a 20–25-m-thick cross-bedded quartzite and black phyllite (Bauer, 1988; Bauer and Helper, 1994). The estimated maximum thicknesses of the Pilar and Piedra Lumbre Formations are

~600 m and 400 m, respectively (Bauer, 1988). The fine-grained, carbonaceous nature of the Pilar Formation suggests deposition in an outer marine shelf (Soegaard and Eriksson, 1986) or deep marine environment.

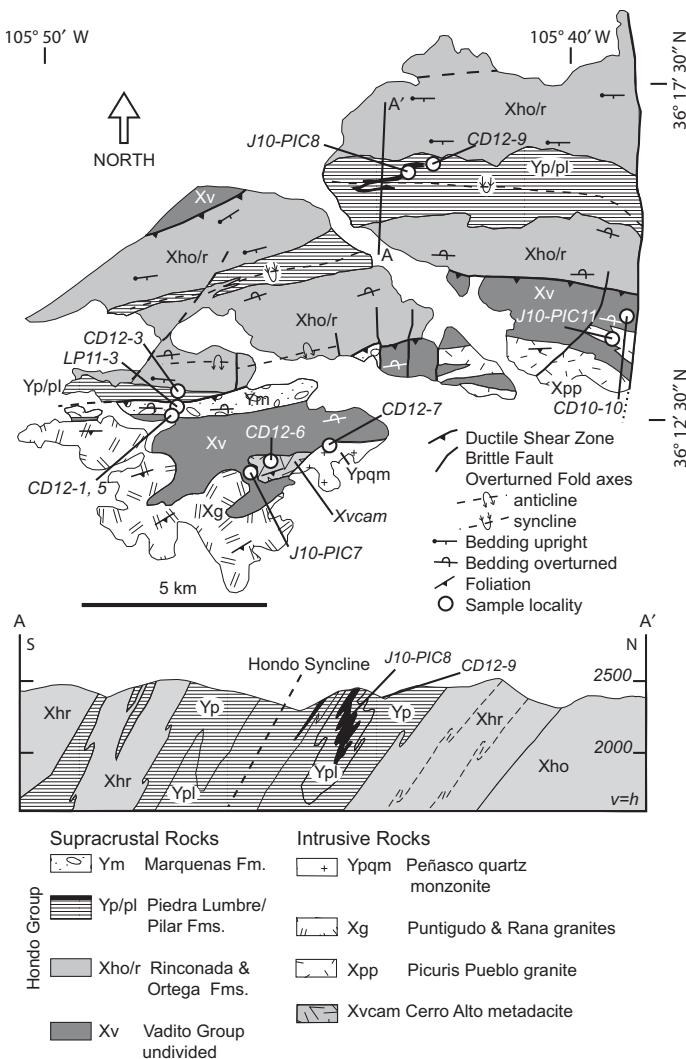
The Mesoproterozoic Marqueñas Formation consists of a basal polymictic boulder to cobble conglomerate overlain by cross-bedded quartzite (Fig. 3) (Soegaard and Eriksson, 1986; Mawer et al., 1990; Bauer, 1993). The quartzite is overlain by a quartz pebble conglomerate, and the entire unit has an apparent thickness of ~500 m. Soegaard and Eriksson (1986) proposed that the Marqueñas Formation had a southern source and was deposited in a braided alluvial plain. Detrital zircon ages give a maximum depositional age of ca. 1457 Ma (Jones et al., 2011). In the Picuris Mountains, the Marqueñas-Vadito Group contact is interpreted to represent an unconformity overprinted by deformation (Jones et al., 2011), and the Marqueñas Formation contact with the Piedra Lumbre Formations is interpreted as a ductile shear zone (Fig. 3; Bauer, 1993).

### Proterozoic Intrusive Rocks

Vadito Group rocks in the southern Picuris Mountains are crosscut by several intrusions, including the Paleoproterozoic Cerro Alto metadacite (ca. 1630 Ma), the Puntiagudo granite porphyry (1684 ± 1 Ma), the Rana quartz monzonite (1673 ± 5 Ma), the Picuris Pueblo granite, and the Mesoproterozoic (ca. 1450 Ma) Peñasco quartz monzonite (Fig. 3) (Long, 1976; Bell, 1985; Bauer, 1988, 1993; Bauer and Helper, 1994). The Paleoproterozoic plutons all contain a moderately to strongly developed foliation and mineral lineation, consistent with post-emplacement deformation. The Mesoproterozoic Peñasco pluton has a strongly foliated and sheared margin with a relatively undeformed interior and is interpreted as a syntectonic pluton (Bauer, 1993). Estimated emplacement depths range from very shallow (<2 km) for the Cerro Alto metadacite, to 1–4 km for the Puntiagudo granite porphyry, to 3–6 km for the Rana quartz monzonite and 8–13 km for the Peñasco quartz monzonite. Notably, no intrusive rocks appear to crosscut the Hondo Group in the region (Bauer and Williams, 1989).

### Timing of Regional Metamorphism and Deformation in Northern New Mexico (Ca. 1650 Ma vs. Ca. 1425 Ma)

Rocks exposed in the Picuris Mountains were part of an east-striking, north-vergent, Proterozoic ductile fold-and-thrust belt that extended across north-central New Mexico (Fig. 2)



**Figure 3.** Simplified geological map and cross section of the Picuris Mountains after Bauer (1988, 1993) showing kilometer-scale folding of Paleoproterozoic and Mesoproterozoic formations, the ductile Pilar and Embudo thrust faults, and approximate sample locations. The solid black unit represents quartzite in the upper Piedra Lumbre Formation.

(Karlstrom and Daniel, 1993; Daniel et al., 1995; Cather et al., 2006). This reconstructed orogenic belt was previously interpreted to record accretion of the Mazatzal Province to southern Laurentia during the ca. 1650 Ma Mazatzal orogeny (Karlstrom and Daniel, 1993; Daniel et al., 1995). Many studies of Proterozoic rocks in New Mexico contend that major folds and penetrative fabrics were formed during the Mazatzal orogeny and were locally reactivated and intensified during the ca. 1400 Ma thermal event (Williams, 1991; Bauer, 1993; Bauer and Williams, 1994; Karlstrom et al., 1997, 2004; Pedrick et al., 1998; Read et al., 1999; Williams et al., 1999). Polyphase metamorphic

pressure-temperature (*P-T*) paths associated with this interpretation are characterized by a ca. 1650 Ma clockwise loop, with cooling at depth followed by isobaric reheating ca. 1400 Ma, implying long-term (~200 m.y.) middle-crustal residence across a broad region (e.g., Williams and Karlstrom, 1996; Pedrick et al., 1998; Read et al., 1999; Karlstrom et al., 2004). Alternatively, Daniel and Pyle (2006) proposed that the dominant regional structures, penetrative deformation, and metamorphism in northern New Mexico were Mesoproterozoic in age, on the basis of ca. 1435–1400 Ma metamorphic monazite grains included within deformed kyanite and sillimanite, and relatively undeformed

andalusite porphyroblasts from the Ortega Formation in the northern Picuris Mountains. This hypothesis was supported by Jones et al. (2011), who reported the first occurrence of Mesoproterozoic detrital zircon within a presumed Paleoproterozoic section in the southwestern United States. Their data constrained the maximum depositional age of deformed, amphibolite-facies metaconglomerate and quartzite of the Marqueñas Formation to be ca. 1450 Ma, some 200 m.y. younger than previously recognized (Bauer and Williams, 1989; Bauer, 1993).

## ZIRCON U-Pb GEOCHRONOLOGY

### Analytical Methods

Zircon U-Pb isotope measurements were conducted at the University of Arizona and at the University of New Brunswick. Zircon mounts for samples J10-PIC7, J10-PIC8, J10-PIC11, CD10-10, and LP11-03 were prepared and analyzed at the University of Arizona LaserChron Center using laser ablation–multicollector–inductively coupled plasma–mass spectrometry (LA-MC-ICP-MS) techniques, as described by Gehrels et al. (2006, 2008). Igneous and detrital zircon samples CD12-1, CD12-3, CD12-5, CD12-6, CD12-7, and CD12-9 were analyzed with laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) at the University of New Brunswick Department of Earth Sciences. Detailed procedures for these analyses are described in the Data Repository.<sup>1</sup> Approximately 110–125 grains were randomly chosen for all detrital zircon samples. We also analyzed an additional 200 grains from sample J10-PIC8 using the LA-ICP-MS facility at the University of New Brunswick to augment the original data set and better define the complex age distribution of this sample. About 50 grains were carefully selected and analyzed for each igneous sample. Grains were characterized and evaluated using transmitted light microscopy, and backscattered electron (BSE) or cathodoluminescence (CL) images prior to and following analysis.

Complete detrital zircon analytical results are presented together with sample location coordinates in the Data Repository (Table DR1 [see footnote 1]), and detrital zircon ages are summarized in Table 1. Data that were >20% discordant or >5% reverse discordant (by comparison of  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{206}\text{Pb}/^{207}\text{Pb}$  ages) were

<sup>1</sup>GSA Data Repository item 2013288, representative cathodoluminescence images of analyzed detrital and igneous zircon, U-Pb concordia diagrams for detrital zircon samples, U-Pb geochronologic methodology, and complete geochronology data, is available at <http://www.geosociety.org/pubs/ft2013.htm> or by request to editing@geosociety.org.

eliminated and are not reported. Complete igneous zircon analytical results are presented together with sample location coordinates in the Data Repository (Table DR2 [see footnote 1]). Uncertainties for individual analyses are reported at  $1\sigma$  or 1 standard error. Uncertainties for calculated mean igneous ages and detrital zircon population ages are reported as  $2\sigma$  or twice the standard error. Representative cathodoluminescence images of detrital zircon grains and igneous zircon grains are included in the Data Repository (Figs. DR1 and DR2 [see footnote 1]). Detrital zircon age distribution diagrams are presented in Figure 4, and corresponding U-Pb concordia diagrams for all detrital analyses are presented in the Data Repository (Fig. DR3 [see footnote 1]). Detrital zircon age probability density plots from this study and selected analyses from Jones et al. (2011) and their approximate lithostratigraphic position within the Vadito and Hondo Groups and the Marqueñas Formation are summarized in Figure 5.

### Detrital Zircon U-Pb Ages

#### Vadito Group

Detrital zircon samples were collected from three different map units within the Vadito Group (Fig. 3). Sample CD12-1 is a quartz-rich schist with compositional layering defined by thin, heavy mineral seams (Fig. 6A); it corresponds to the Vs5 map unit of Bauer and Helper (1994) and was collected 1–2 m below the disconformable contact with the basal Marqueñas Formation metaconglomerate. Sample CD10-10 is a feldspathic quartzite with well-defined, overturned, cross-beds that are younger to the north (Fig. 6B). This particular quartzite unit was previously mapped as the Marqueñas Formation by Bauer and Helper (1994), but it is shown as Vadito Group quartzite by Bauer (1988). A detrital zircon sample was collected, in part to test the stratigraphic association of the map unit. Sample J10-PIC7 is a quartz pebble metaconglomerate from the lower Vadito Group quartzite (unit Vq of Bauer and Helper, 1994). Cross-beds in this unit are overturned and face north. The quartzite and metaconglomerate layers are crosscut by the  $1673 \pm 5$  Ma Rana pluton (Bell, 1985), thus providing a lower age bracket on Vadito Group deposition. Detrital zircon from the three Vadito Group samples yielded ages ranging from 2557 Ma to 1641 Ma (Table 1; Data Repository Table DR1 [see footnote 1]; Figs. 4A–4C), and all but four grains out of 305 analyzed were Paleoproterozoic. Age probability distribution diagrams for each sample show a strong unimodal population characterized by narrow peaks with ages of 1715 Ma, 1705 Ma, and 1707 Ma (Table 1; Figs. 4A–4C).

TABLE 1. DETRITAL ZIRCON AGE SUMMARY

		$N_{\text{total}}$	Age range (Ma)	Peak age ranges (Ma)			Age peak(s) (Ma)	$N_{\text{peak}}$	Age peak(s) (Ma)	$N_{\text{peak}}$
<b>Marqueñas Formation</b>										
LP11-3	99	3530–1459	1450	1495	4	1471	4	1799	9	
			1609	1833	87	1697	51	1824	4	
			1836	1884	2	1711	56	1862	3	
			1906	1919	2	1779	13	1912	3	
CD12-5	120	2786–1417	1419	1511	8	1472	8			
			1616	1819	97	1716	51			
			1844	1876	4	1792	18			
						1859	4			
<b>Upper Hondo Group</b>										
<i>Piedra Lumbre Formation</i>										
J10-PIC8	244	2698–1451	1446	1894	166	1475	15	1789	29	
						1510	24	1830	14	
						1550	31	1852	15	
						1592	43	1903	7	
						1611	51	1994	4	
						1728	38			
<i>Rinconada Formation</i>										
CD12-3	105	2732–1651	1641	2011	93	1679	12	1910	9	
						1732	18	1955	9	
						1759	24	2008	4	
						1853	12			
<i>Vadito Group</i>										
CD10-10	93	1772–1691	1650	1766	92	1715	88			
CD12-1	116	2553–1641	1650	1890	111	1705	77	1856	3	
J10-PIC7	95	2557–1677	1598	1923	94	1707	93			

*Note:* Only 80%–105% concordant analyses are reported and summarized.  $N_{\text{total}}$  refers to total number of grains included in the age pick analysis. Age range is the total range of ages of all grains included in the age pick analysis. Peak age range represents the range of age-probability contributions (at  $2\sigma$ ) from three or more analyses.  $N_{\text{range}}$  indicates number of ages that contributed to the peak age range. Age peaks represent maxima in the age probability curve consisting of three analyses or more.  $N_{\text{peak}}$  indicates the number of analyses that contribute age probability to an age peak. Age pick data were calculated using Microsoft Excel macros made available by G. Gehrels at the University of Arizona LaserChron Center (Gehrels, 2009).

#### Rinconada and Piedra Lumbre Formations

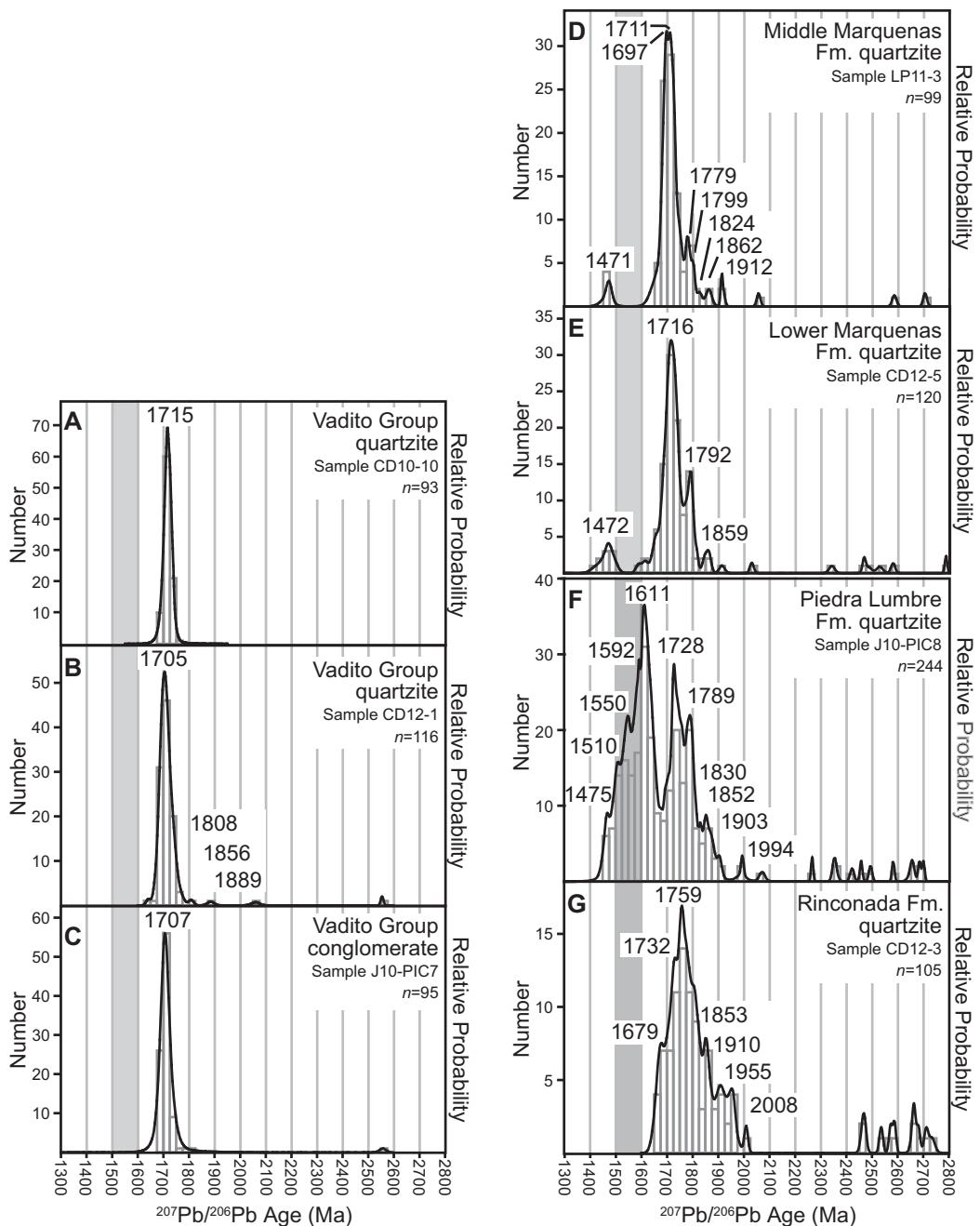
Two new detrital zircon samples were collected from the upper Hondo Group (Fig. 3) to evaluate provenance patterns through the top of the section relative to data previously published for the Ortega and Rinconada Formations by Jones et al. (2011). One sample (CD12-3) was collected from black, vitreous quartzite (Fig. 6C) at the top of the Rinconada Formation on the southern limb of the Copper Hill anticline. Several discontinuous layers of this distinctive unit were observed in outcrop, and we interpret this pattern to represent small-scale fault repetition on the basis of an observed abundance of surfaces containing slickenlines. Alternatively, the multiple layers may represent distinct beds within a mixed sandstone-shale protolith. Micaeous quartzite (J10-PIC8; Fig. 6D) was collected from the upper Piedra Lumbre Formation near the hinge of the Hondo syncline and near the top of the exposed section.

Age probability diagrams for the two samples (Figs. 4F and 4G) show more diverse detrital zircon age distributions relative to the underlying parts of the stratigraphy. Zircon age ranges were 2732–1651 Ma and 2698–1451 Ma in the Rinconada and Piedra Lumbre samples, respec-

tively (Table 1). The Rinconada and Piedra Lumbre Formations yielded eight and five Archean grains, respectively. Seven statistically significant Proterozoic age populations in the Rinconada Formation range from 2008 Ma to 1679 Ma, with the dominant age probability peak at 1759 Ma (Fig. 4G). The youngest population gives a maximum depositional age of 1679  $\pm$  18 Ma for the unit ( $n = 12$ ; Table 1). Eleven age populations in the Piedra Lumbre Formation sample range from 1994 Ma to 1475 Ma, with a dominant peak at ca. 1611 Ma (Fig. 4F). Seventy-four Mesoproterozoic detrital zircons with ages ranging from 1598 Ma to 1451 Ma make up 30% of the total population and define significant peaks at 1592 Ma, 1550 Ma, 1510 Ma, and 1475 Ma (Table 1). The youngest population gives a maximum depositional age of 1475  $\pm$  19 Ma ( $n = 15$ , mean square of weighted deviates [MSWD] = 1.7; Table 1).

#### Marqueñas Formation

Two new samples from the Marqueñas Formation (Fig. 3) were collected to complement and to confirm detrital zircon results previously reported by Jones et al. (2011). One sample (LP11-3) was collected from the middle quartzite member



**Figure 4. Complete age distribution diagrams for  $^{207}\text{Pb}/^{206}\text{Pb}$  ages from samples analyzed in this study. Plots show all peak ages defined by the occurrence of at least three grains yielding identical ages within uncertainty. Histogram bins are 25 m.y. wide. The vertical gray bar corresponds to the Laurentian “magmatic gap” time period between 1600 and 1500 Ma. Corresponding concordia diagrams are in the Data Repository (Fig. DR3 [see text footnote 1]).**

(Fig. 6E) ~400 m to the east, along strike from sample J07-PIC2 reported by Jones et al. (2011). We collected a second sample (CD12-5) from a coarse sand to granule layer within the basal polymictic cobble to boulder metaconglomerate (Fig. 6F) exposed in a road cut along State Highway 75. This sample was ~3 m above the Vadito Group–Marqueñas Formation unconformity and represents the lowermost sample from the Marqueñas Formation. Quartzite from the middle Marqueñas Formation (LP11-3) yielded ages ranging from 3530 Ma to 1459 Ma, with only a few percent of Archean grains (Table 1; Fig.

4D). Paleoproterozoic detrital zircons are dominant, defining five subsidiary populations from 1912 Ma to 1779 Ma and a split maximum peak with ages of 1711 Ma and 1697 Ma (Fig. 4D). A minor but statistically significant population of Mesoproterozoic grains yielded a mean age of  $1471 \pm 20$  Ma ( $n = 4$ , MSWD = 0.2; Table 1). The quartzite lens from within the basal conglomerate (CD12-5) yielded a similar range of grain and population ages (Fig. 4E). In total, 120 grains with ages between 2786 Ma and 1417 Ma define a maximum age probability peak at 1716 Ma and subsidiary peaks at 1792 Ma and 1859 Ma

(Table 1). A small population of Mesoproterozoic grains defines a peak at  $1472 \pm 18$  Ma ( $n = 8$ , MSWD = 1.2; Table 1), thus providing maximum depositional ages for the lower and middle parts of the Marqueñas Formation that are identical within uncertainty.

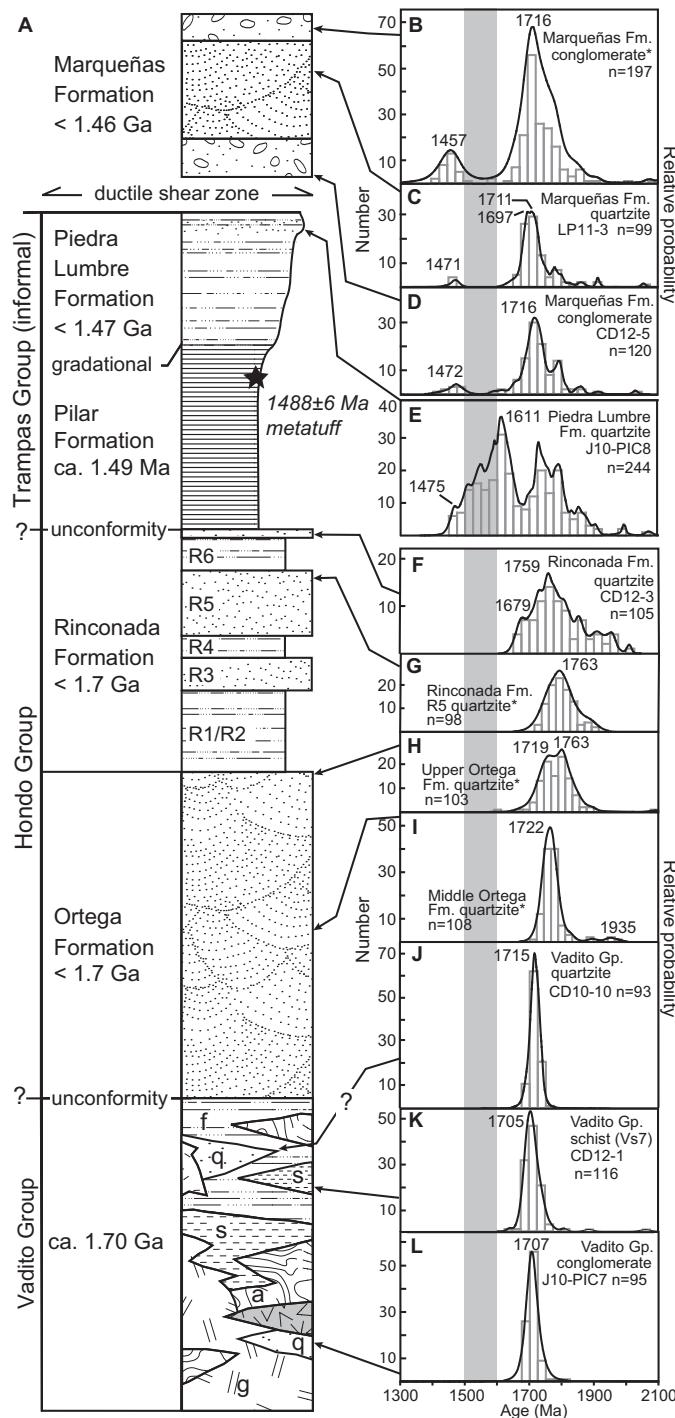
#### Igneous Zircon U-Pb Ages

We sampled four igneous units that crosscut or are interlayered with the Vadito and Hondo Groups to better constrain the ages of deposition, deformation, and metamorphism in the Picuris

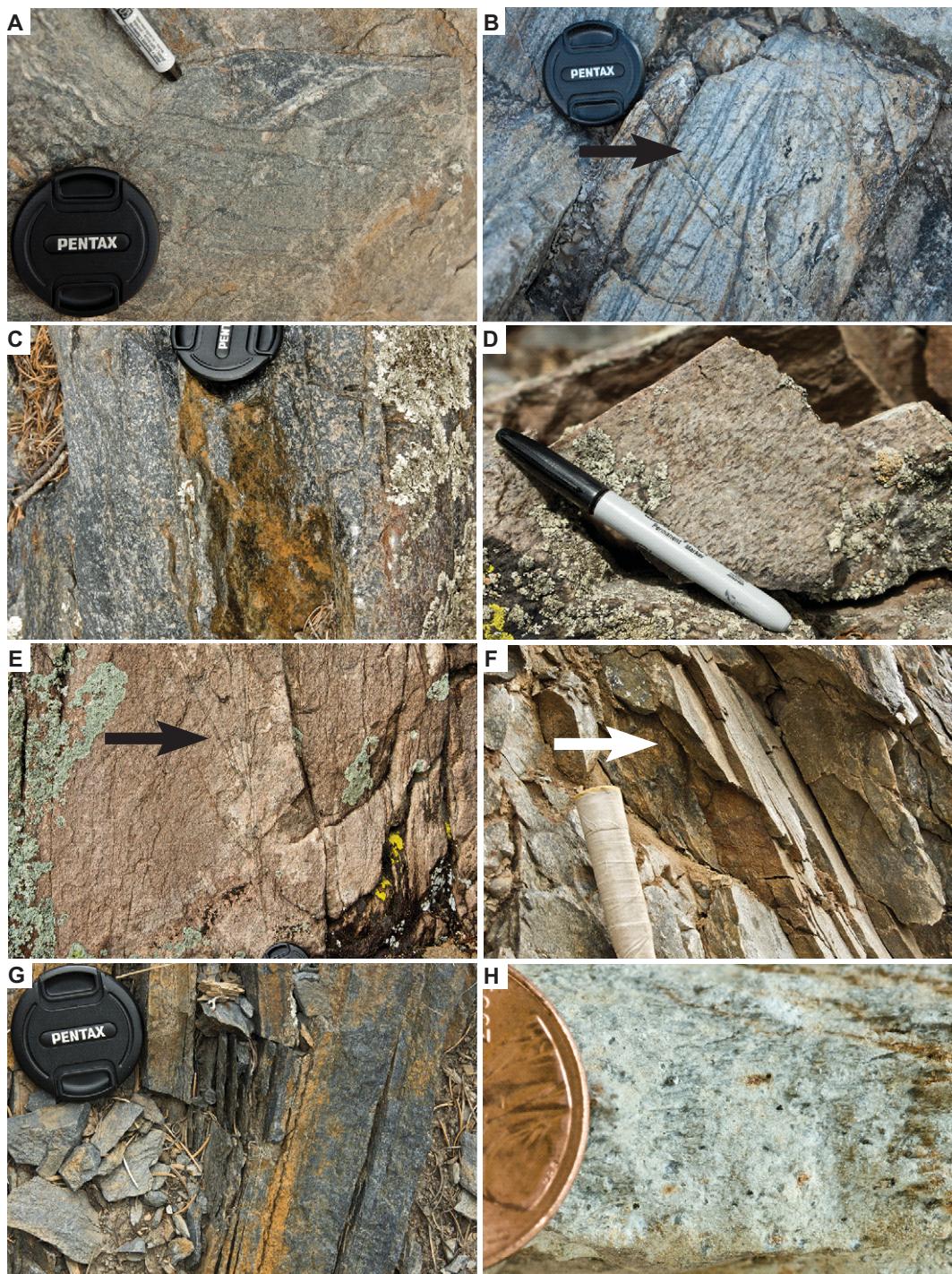
Mountains. Three intrusive units were sampled, including the Cerro Alto metadacite (CD12-6), the Picuris Pueblo granite (J10-PIC11), and the Peñasco quartz monzonite (CD12-7). The Cerro Alto metadacite is generally pervasively foliated, but we sampled lower-strain domains within the pluton that were characterized by subhedral to euhedral feldspar phenocrysts and a relatively lower degree of foliation intensity. Zircon from the Cerro Alto metadacite were euhedral to subhedral and prismatic with pyramidal terminations. Grains ranged in size from ~100  $\mu\text{m}$  to 250  $\mu\text{m}$ . Cathodoluminescence (CL) imaging revealed concentric internal zonation that is characteristic of igneous crystallization (Fig. DR2 [see footnote 1]). Several grains contain dark cores with truncated growth zones that we interpret to represent xenocrystic cores. Forty-eight grains analyzed yielded ages ranging from 2032 to 1682 Ma (Table DR2 [see footnote 1]). Twenty grains defined a coherent group of ages with a mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1710 \pm 10$  Ma (MSWD = 1.6; Fig. 7A). We interpret this age to represent crystallization of the dacite and attribute the older ages to inheritance.

We collected one sample of coarse-grained biotite granite from a series of small and strongly weathered exposures of the Picuris Pueblo granite (J10-PIC11). The granite appeared to intrude thin slivers of quartzite that are correlative with feldspathic quartzite described earlier herein from the Vadito Group (sample CD10-10). This sample contained subhedral to euhedral, prismatic zircon 100–300  $\mu\text{m}$  in length. CL images reveal concentric internal growth zones without any observed evidence for inherited cores (Fig. DR2B [see footnote 1]). All 24 grains analyzed from this sample defined a coherent group with an age range of 1707–1684 Ma and a mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1699 \pm 3$  Ma (MSWD = 0.4; Table DR2 [see footnote 1]; Fig. 7C).

A sample of the Peñasco quartz monzonite was collected from the relatively undeformed, megacrystic interior ~30 m from the deformed margin of the pluton (CD12-7). Outcrops are well exposed but rounded with moderate weathering. We avoided the relatively abundant ellipsoidal mafic enclaves during sample collection. Zircon from this sample was subhedral to euhedral, generally prismatic and elongate, and 100–200  $\mu\text{m}$  in size. CL images reveal concentric internal growth zonation but also some irregular internal regions with truncated zonation or no zoning (Fig. DR2C [see footnote 1]). We interpret these interior regions to represent xenocrystic cores. Fifty-one grains analyzed had ages ranging from 1683 to 1408 Ma (Table DR2 [see footnote 1]), and 32 grains defined a coherent group with a mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1450 \pm 10$  Ma (MSWD = 1.04; Table DR2 [see foot-



**Figure 5.** (A) Generalized composite lithostratigraphic section for the Mesoproterozoic Marqueñas Formation and the Hondo and Vadito Groups for northern New Mexico (f—felsic schist, a—amphibolite, q—quartzite with minor conglomerate, s—schist, g—granitic intrusion). (B–L) Age distribution diagrams for  $^{207}\text{Pb}/^{206}\text{Pb}$  ages including new data from this study (C, D, E, F, J, K, and L) and previously published detrital zircon data for the Ortega and Rinconada Formations (B, G, H, and I) from Jones et al. (2011). Plots show selected peak ages defined by the occurrence of at least three grains yielding identical ages within uncertainty. Histogram bins are 25 m.y. wide. The vertical gray bar corresponds to the “magmatic gap” time period between 1600 and 1500 Ma. The star indicates the approximate location of the 1448 Ma meta-tuff within the Pilar Formation. See text for discussion. Asterisks indicate data for plots taken from Jones et al., 2011.



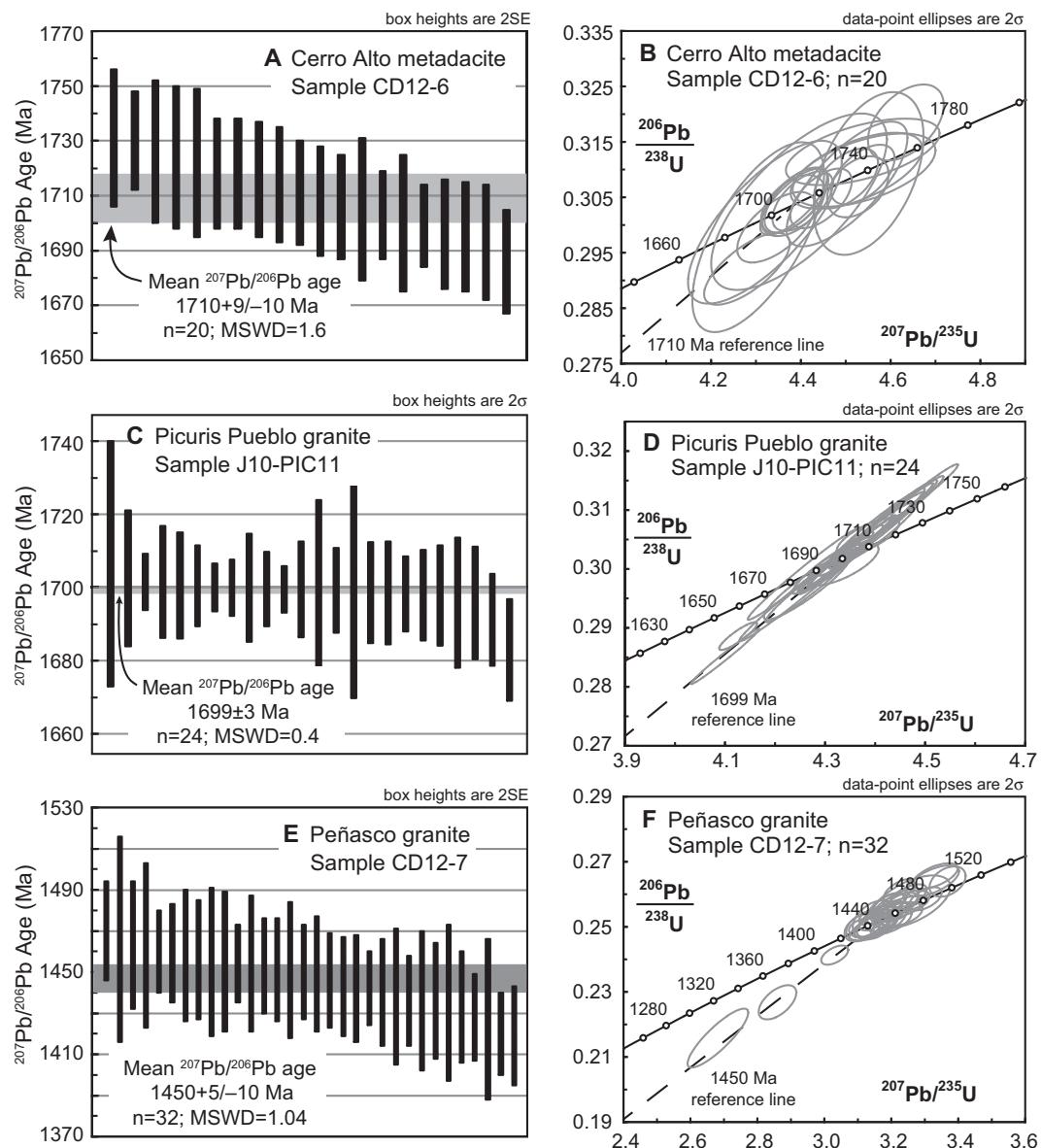
**Figure 6.** Outcrop photographs showing detrital zircon samples used in this study. (A) Sample CD12-1, schist from the uppermost Vadito Group with poorly defined compositional layers that may be relict cross-beds. (B) Sample CD10-10, overturned, north-facing cross-bedded quartzite from the Vadito Group in southeastern Picuris. (C) Sample CD12-3, black, vitreous quartzite from the basal uppermost Rinconada Formation. (D) Sample J10-PIC8, micaceous quartzite from the uppermost Piedra Lumbre Formation. (E) Sample LP11-3, overturned, north-facing cross-bedded quartzite from the middle Marqueñas Formation. (F) Sample CD12-5, coarse sand to granule layer in the lowermost basal conglomerate of the Marqueñas Formation. Hammer handle for scale. (G) Representative outcrop of black, fine-grained Pilar Formation. (H) Close-up image of sample CD12-9, interpreted as a metamorphosed tuff layer within the upper Pilar Formation.

note 1]; Fig. 7E). We interpret the older ages to represent inheritance and three younger ages to represent partial Pb loss, perhaps during subsequent metamorphism (Daniel and Pyle, 2006).

We also sampled one of several centimeter-to-meter-scale white schistose layers within the Pilar Formation (Long, 1976; Bauer, 1988). These layers are particularly distinctive within the black, carbonaceous phyllite (Figs. 6G and

6H), and we interpret these layers to represent metamorphosed tuffs on the basis of their white color, fine grain size, microporphroblasts of microcline, and infrequent interlayered occurrence within the otherwise homogeneous, fine-grained sedimentary succession. We sampled a 1–2-m-thick meta-tuff (CD12-9) from the northern limb of the Hondo syncline in the northern Picuris Mountains (Fig. 4A). The sample

yielded abundant small, euhedral zircon with an average size of ~50  $\mu\text{m}$ . Cathodoluminescence images reveal concentric igneous growth zones with some more irregular interior regions and truncated zones, but all of the grains are igneous in appearance, consistent with derivation from a tuff layer (Fig. 8A). In total, 123 grains analyzed yielded an age range of 1701–1407 Ma (Table DR2 [see footnote 1]), and 102 of these



**Figure 7. Mean weighted age and concordia plots for (A, B) igneous samples from the Cerro Alto metadacite, (C, D) Picuris Pueblo granite, and (E, F) the Peñasco quartz monzonite. See text for discussion. MSWD—mean square of weighted deviates.**

grains defined a coherent group with a mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1488 \pm 6$  Ma (MSWD = 1.1; Fig. 8B). We interpret this age to represent the depositional age of the volcanic protolith.

## INTERPRETATION AND DISCUSSION

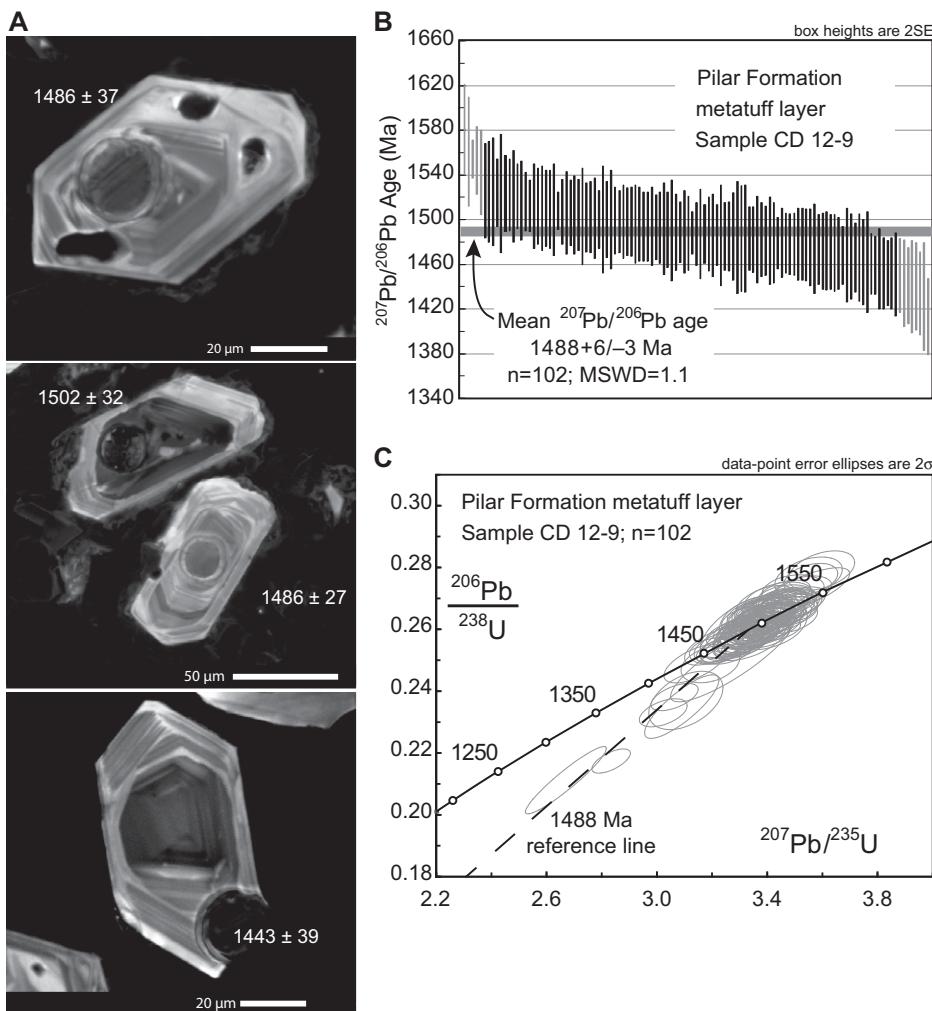
### Age and Provenance of the Vadito Group

Detrital zircon age populations from the Vadito Group overlap in age with the 1710 Ma Cerro Alto metadacite and the 1699 Ma Picuris Pueblo granite. The southern Vadito conglomerate (J10-PIC7) has a peak age of 1707 Ma identical within uncertainty to the age of the crosscutting Cerro Alto metadacite, and field relations clearly show that the Cerro Alto meta-

dacite intrudes the lower Vadito Group. No intrusions appear to crosscut the upper Vadito Group schist (CD12-1). Contact relationships between the eastern Vadito quartzite (CD10-10) and the Picuris Pueblo granite are less clear. Although Bauer (1988) interpreted the Picuris Pueblo granite to be intrusive, he noted that poor exposure, significant weathering, and penetrative deformation might have masked a folded and/or faulted unconformity between the eastern quartzite and the Picuris Pueblo granite. The feldspathic nature of the sample suggests local derivation possibly from the Picuris Pueblo granite. The stratigraphic position of this unit relative to the Vadito schist sample is not clear, but we tentatively place it higher in the section because the Plomo fault appears to cut up sec-

tion in the Vadito Group exposures from west to east. Overall, the measured detrital zircon age distributions and the observed crosscutting relationships are consistent with the interpreted ca. 1710–1700 Ma age for the Vadito Group given by Bauer and Williams (1989).

In comparison with published data from Vadito Group exposures in the Tusas Mountains to the west (Jones et al., 2011), the narrow unimodal age spectra for the Picuris Mountains samples suggest more restricted local sources that likely included ca. 1700 Ma meta-rhyolite and possibly metadacite or granite intruding the lower parts of the Vadito Group or ca. 1700 Ma granitic plutons exposed in the Cimarron and Rincon Mountains (Bauer and Williams, 1989; Pedrick et al., 1998; Read et al., 1999). The similarity of



**Figure 8.** (A) Representative cathodoluminescence images of zircon recovered from the Pilar Formation meta-tuff. Euhedral crystals show little evidence of sedimentary reworking, and zoning is characteristic of igneous crystallization. Ages are in Ma. (B) Weighted mean age plot for zircon recovered from the meta-tuff gives a crystallization age of  $1488 \pm 6$  Ma. (C) Concordia plot for zircon from the meta-tuff. MSWD—mean square of weighted deviates.

detrital zircon age spectra for all of the Vadito Group samples (Figs. 5J–5L) and the lack of Mesoproterozoic detrital zircons in the eastern quartzite (CD10–10) are consistent with placing it into the Vadito Group and not the Marqueñas Formation as shown by Bauer and Helper (1994). The close association between Vadito Group detrital zircon ages and the age of the Cerro Alto metadacite and Picuris Pueblo granite suggests contemporaneous sedimentation and magmatism in an active tectonic environment such as a continental rift (Mawer et al., 1990), or intra-arc or back-arc setting (Soegaard and Eriksson, 1986) associated with the waning stages of the Yavapai orogeny. Similar intra-arc to back-arc settings have been proposed for the coeval upper portion of the Tonto Basin Supergroup in Arizona (Condie et al., 1992; Cox et al., 2002).

## Age and Provenance of the Hondo Group

### Lower Hondo Group

The peak shapes for the Ortega and Rinconada Formations of the lower Hondo Group (Figs. 5G–5I; Jones et al., 2011) are similar to the underlying Vadito Group. However, the probability density curves become broader up section, and the peak ages are systematically older higher in the section. We interpret these patterns to reflect local unroofing of older arc-related igneous basement complexes such as the Pecos and Gold Hill complexes and/or shifts to older source regions in the Yavapai Province through time. This interpretation is supported by the geochemical study of McLennan et al. (1995), who showed that ca. 1700 Ma Nd-model ages and Pb-isotope ratios within the Ortega and

Rinconada Formations are essentially identical with values from nearby mafic metavolcanic complexes and plutons that yield crystallization ages and Nd-model ages near 1700 Ma (DePaolo et al., 1991). McLennan et al. (1995) concluded that a major contributor of sediment to the quartzites and pelites in the Hondo Group was a low Th/U Paleoproterozoic terrane such as the southern Colorado–northern New Mexico Pb-isotope crustal province (Aleinikoff et al., 1993) and that there was minor contribution (10% on average) from Archean sources, possibly the Wyoming Province (Fig. 9B).

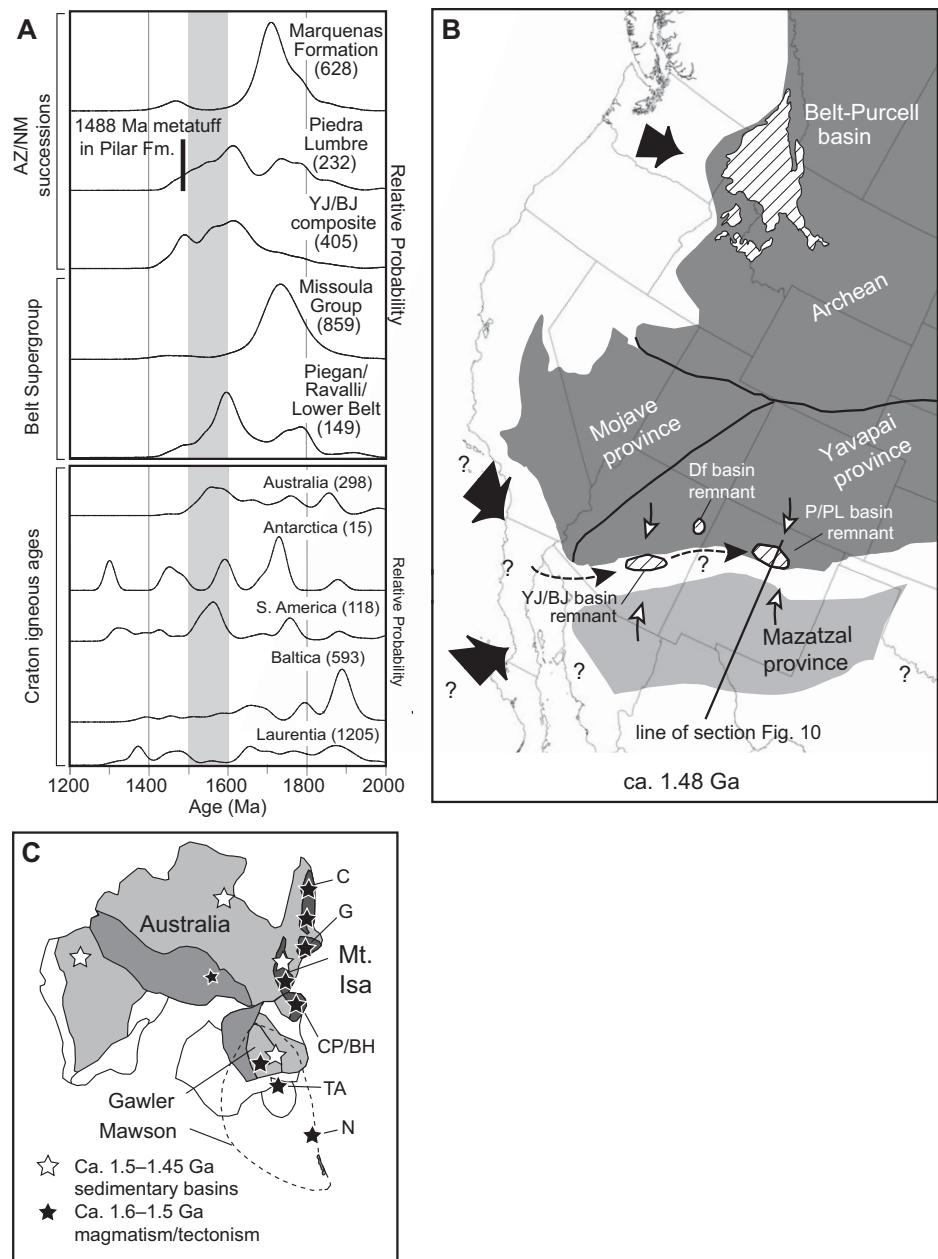
The 1–1.5-km-thick Ortega Formation was interpreted by several previous studies to be deposited conformably with the Vadito Group ca. 1700 Ma (Soegaard and Eriksson, 1985, 1986; Mawer et al., 1990; Bauer, 1993; Bauer and Williams, 1989), and there is no obvious angular discordance across the Vadito Group–Hondo Group contact (Grambling and Coddington, 1982; Bauer and Williams, 1989). In the Tusas Mountains, the contact is locally described as gradational (Bauer and Williams, 1989; Williams, 1991). However, there is no evidence that precludes the possibility that the contact is a disconformity. Unlike the Vadito Group, which is crosscut by Paleoproterozoic and Mesoproterozoic intrusions, no intrusive rocks unequivocally intrude the Hondo Group. The best available age constraints on the lower Hondo Group include ca. 1723 Ma detrital zircon age populations (Jones et al., 2011) and ca. 1434 Ma metamorphic monazite (Daniel and Pyle, 2006), permissive of a wide range of depositional ages.

### Trampas Group (Formerly Part of Hondo Group)

The Pilar and Piedra Lumbre Formations of the upper Hondo Group were previously interpreted to be Paleoproterozoic and conformable with the lower Hondo Group (Soegaard and Eriksson, 1985, 1986; Bauer and Williams, 1989; Bauer, 1993). However, our findings require that both units were deposited during the Mesoproterozoic. The age of the interlayered meta-tuff unit in the upper Pilar Formation indicates that it was deposited at  $1488 \pm 6$  Ma, consistent with a maximum depositional age of ca. 1475 Ma in the overlying Piedra Lumbre Formation as indicated by detrital zircon ages.

The eruptive center for the metamorphosed tuff is not obvious, and plutons of that age are not common in the western United States. One possible candidate is the ca. 1480 Ma (no reported uncertainty) Macho Creek granite that intrudes the Jones rhyolite complex in the Pecos complex (Robertson and Condie, 1989). Plutons within this age range in Colorado include

**Figure 9.** (A) Normalized relative probability density distribution curves for detrital zircon from this study, the Hess Canyon Group of Arizona (Doe et al., 2012), the Belt Supergroup (Ross and Villeneuve, 2003; Link et al., 2007), and composite igneous zircon age spectra from candidate source cratons (Condie et al., 2009; Goodge et al., 2010; Chemale et al., 2011; Ibanez-Mejia et al., 2011; Teixeira et al., 2013). See text for discussion. YJ/BJ—Yankee Joe/Blackjack formations. (B) Speculative paleogeographic reconstruction showing the relative locations of ca. 1490–1460 Ma basins and a possible depositional system for detrital zircon to travel from the Laurentian margin across the southwestern United States. Axial basin transport moves detritus from highlands in the west-southwest across Arizona (AZ) and into New Mexico (NM). Additional sediment influx may originate from the Yavapai and possibly the Mazatzal cratonic provinces to the north and south, respectively. Approximate line of section for Figure 10 is shown. (C) Stars indicate areas of ca. 1650–1500 Ma magmatism/tectonism on the Australian/Antarctic craton. Dashed outline corresponds to the Mawson continent. C—Coen Inlier; CP/BH—Curnamona Province/Broken Hill subdomain; G—Georgetown Inlier; N—Nimrod; TA—Terra Adelie craton. Remnants of Mesoproterozoic basins, Yankee Joe/Blackjack (YJ/BJ), Defiance (Df), Pilar/Piedra Lumbre (P/PL) after Doe et al., 2013, and this work.



the  $1486 \pm 36$  Ma granite of Williams Creek and the  $1474 \pm 7$  Ma McCoy Gulch stock (Bickford et al., 1989). Circa 1470–1460 Ma granitic plutons in the Burro Mountains of southwestern New Mexico are apparently too young to be a potential source for the meta-tuff (Amato et al., 2011). However, two plutons from the Caballo Mountains in southern New Mexico yield zircon crystallization ages of  $1486 \pm 16$  Ma and  $1487 \pm 24$  Ma (Amato and Becker, 2012), identical in age with the meta-tuff. Precambrian rocks in the St. Francois Mountains, ~1400 km to the east in Missouri, are interpreted to represent the remains of a ca. 1480–1400 Ma caldera complex (Sides et al., 1981; Van Schmus et al.,

1996; Menegh et al., 2002) and could be a possible distal source for the meta-tuff.

We interpret the contact between the Rinconada and Pilar Formations to be a disconformity. It is also possible that a significant disconformity exists across the contact between the Vadito and Hondo Groups. Together these two proposed unconformities could account for a hiatus of 200 m.y. or more, and yet there is no obvious evidence for disparate deformation histories across the section (Nielson and Scott, 1979; Holcombe and Callender, 1982; Bauer, 1993). We place the Pilar-Rinconada unconformity above the distinctive black quartzite (CD12-3) in the lower Pilar Formation, although

it could possibly be at the base of the quartzite. The transition from deltaic deposits of the Rinconada Formation to deep marine or outer shelf deposits of the Pilar Formation (Soegaard and Eriksson, 1986) is compatible with a model involving flooding of the Paleoproterozoic succession during basin formation or reactivation and renewed deposition ca. 1490 Ma or later. On the basis of our new data, we suggest that the Pilar and Piedra Lumbre Formations be removed from the Hondo Group. Given the apparently conformable nature of the two formations (Bauer, 1988), designation as a group is appropriate, and we informally refer to these two formations as the Trampas group.

Geochemical data from the Pilar and Piedra Lumbre Formations are somewhat different relative to the Hondo Group rocks (McLennan *et al.*, 1995). Several samples from the Pilar and Piedra Lumbre Formations showed significantly lower values for Cr, Sc, Co, Rb, Th, and Sn but were enriched in U, Mo, and V, giving low Cr/V and Th/U values relative to the Ortega and Rinconada Formations. McLennan *et al.* (1995) interpreted these anomalously low values to be related to diagenetic processes. These samples also yielded anomalous rare earth element (REE) patterns that showed light (L) REE depletion and heavy (H) REE enrichment relative to samples from the lower Hondo Group, suggesting that these rocks experienced sedimentation processes similar to those experienced by modern black shales (McLennan *et al.*, 1995). Notably, the samples with anomalous REE patterns typically yielded high  $^{206}\text{Pb}/^{204}\text{Pb}$  and defined an evolution line with an age of  $1430 \pm 25$  Ma, suggesting that the age of sedimentation was significantly younger than ca. 1700 Ma (McLennan *et al.*, 1995), consistent with our findings.

The detrital zircon age spectra suggest a slight change in provenance during this time as well. The dominant detrital zircon population is approximately the same age from the Rinconada Formation to the basal Pilar Formation, suggesting that the primary source areas remained the same after the inferred hiatus. However, the uppermost Rinconada Formation contains a distinct 1679 Ma population that is absent in the lower Hondo Group and in all other ca. 1700 Ma successions in the surrounding region (Jones *et al.*, 2009, 2011). The age of the young population matches peak ages in multiple metasedimentary successions in central and southern New Mexico that were deposited ca. 1660–1600 Ma and derived from local Mazatzal sources (e.g., Amato *et al.*, 2008). The minimum age peak is also similar to the age of the Puntiagudo granite porphyry (ca. 1680 Ma) and the Rana granite (ca. 1673 Ma) from the southern Picuris Mountains. The presence of the younger population is distinct relative to the older, underlying units, and the younger than 1700 Ma grains likely reflect either local igneous sources that intruded the Yavapai Province, or sources in the Mazatzal Province to the south.

In addition to the unexpected ca. 1475 Ma detrital zircon population, the Piedra Lumbre Formation contains the broadest distribution of detrital zircon ages among all samples in the region (Jones *et al.*, 2011). We interpret age populations of 1700 Ma and older to represent a mixture of Yavapai-aged sources and older sources such as the Trans-Hudson orogen or Black Hills to the north (Van Schmus *et al.*, 1987; Dahl *et al.*, 2006). Grains of this age might also reflect older Yavapai and Mojave Province sources to the

west and northwest or possibly detritus from Australia or Antarctica (Boger, 2011).

The dominant ca. 1611 Ma detrital zircon population in the Piedra Lumbre Formation is substantially younger than the 1679 Ma population in the underlying Rinconada Formation. Although igneous rocks as young as ca. 1601 Ma have been reported from the Mazatzal Province (Luther *et al.*, 2006; Luther, 2006), rocks younger than ca. 1630 Ma are not common (Karlstrom *et al.*, 2004; Amato *et al.*, 2008). Furthermore, zircon ages between 1598 and 1475 Ma are not common in all of southern Laurentia (Condie *et al.*, 2009; Voice *et al.*, 2011). Circa 1611 Ma detrital zircon could represent local sources in the Mazatzal Province (Fig. 9B). However, given their relatively high abundance compared to older Paleoproterozoic detritus and a continuous spectrum of detrital zircon ages between 1629 and 1495 Ma, we interpret these age populations to reflect exotic, non-Laurentian sources. Zircons in the youngest population, ca. 1480–1470 Ma, are interpreted to be more locally derived, possibly from areas such as the Macho Creek pluton in the Pecos complex (Robertson and Condie, 1989; Melis, 2001), the Caballo and Burro Mountains in southern New Mexico (Amato *et al.*, 2008, 2011; Amato and Becker, 2012), or their eruptive equivalents.

### Age and Provenance of the Marqueñas Formation

Detrital zircon age populations from two samples of the Marqueñas Formation are similar to each other (Figs. 5B–5C) and to data previously published by Jones *et al.* (2011). We interpret the dominant age populations to reflect input from proximal Yavapai- and Mazatzal-aged sources derived from a southern source area (Soegaard and Eriksson, 1986). We also note the relatively restricted range of Paleoproterozoic age populations, compared to the broader age distribution in the Trampas group, and interpret this pattern to reflect a predominance of local Yavapai-aged sources rather than older sources in the more distal portions of the province to the north (Fig. 9B). Both samples contained a statistically significant component of Mesoproterozoic detrital zircon, and the ages of the populations are identical within uncertainty. Detrital zircon from quartzite within the apparently lowermost section of the basal conglomerate gives a maximum depositional age of  $1472 \pm 18$  Ma (Fig. 4E), and pebble conglomerate from the uppermost member gives a maximum depositional age of  $1453 \pm 10$  Ma (Jones *et al.*, 2011). We interpret the youngest age population to be locally derived from ca. 1450–1480 Ma plutons in the region or their eruptive equivalents.

The youngest detrital zircon peak age for the Marqueñas Formation is ca. 1457 Ma, identical within uncertainty to the oldest reported metamorphic mineral ages in the Picuris Mountains of  $1461 \pm 13$  Ma (Lanzirotti and Hanson, 1997) and  $1456 \pm 16$  Ma (Aronoff *et al.*, 2012) for garnet from the Rinconada Formation and Vadito Group schist, respectively, and  $1454 \pm 7$  Ma for staurolite from the Rinconada Formation (Lanzirotti and Hanson, 1997). The overlap in ages between the maximum depositional age of the Marqueñas Formation and metamorphic mineral ages from the Vadito Group schists and the Rinconada Formation is consistent with a history of deposition closely followed by thrust burial, ductile deformation, and regional metamorphism.

### Provenance of Circa 1600–1500 Ma Detrital Zircon

Detrital zircon populations between 1600 Ma and 1500 Ma (62 of 244 grains) in the Piedra Lumbre Formation represent sources that are uncommon in western Laurentia and are also relatively rare worldwide (Fig. 9A; Condie *et al.*, 2009; Voice *et al.*, 2011). Our discovery marks the second Mesoproterozoic metasedimentary succession in the southwest United States in which 1600–1500 Ma detrital zircon are recognized. Doe *et al.* (2012) reported 1600–1500 Ma detrital zircons in quartzite in the Yankee Joe and Blackjack Formations of the upper Hess Canyon Group in central Arizona. Age constraints indicate that these two units were deposited between 1488 and 1436 Ma, which overlaps with the depositional age bracket of the Pilar and Piedra Lumbre Formations. Statistical comparisons of the Mesoproterozoic detrital zircon ages from the Piedra Lumbre Formation with both Arizona units reveal significant similarity between multiple samples, indicating that source areas were not statistically different for all but the uppermost and lowermost parts of the Arizona succession (Table 2). Thus, we speculate that these two basins represent broadly contemporaneous systems that were both connected to the same exotic sources between ca. 1490 Ma and 1460 Ma.

Detrital zircons between 1600 and 1500 Ma are also reported in multiple units of the Mesoproterozoic Belt Supergroup (Fig. 9A), a >15-km-thick succession of predominately siliciclastic strata exposed in the northwest United States and southwestern Canada. Ross *et al.* (1992) first reported 1600–1500 Ma detrital zircon ages from Lower Belt and Ravalli Group exposures from Montana, Idaho, and southern British Columbia. Link *et al.* (2007) found detrital zircon ages clustered at 1590 Ma

TABLE 2. COMPARISON OF DETRITAL ZIRCON AGES USING THE K-S STATISTIC

K-S statistic		Upper Hess Canyon Group, Arizona				Picuris Mountains, New Mexico			Belt Supergroup	
Sample		YJ00208*	YJ00108*	BJMFD08*	BJ00208*	BJ2-2011*	Piedra Lumbre	Marquenas	Lower Belt composite†	Upper Belt composite†
YJ00108*		0.000								
BJMFD08*		0.000	<b>1.000</b>							
BJ00208*		0.000	<b>0.981</b>	<b>0.985</b>						
BJ2-2011*		0.000	0.002	0.002	0.000					
Piedra Lumbre		0.000	<b>0.357</b>	<b>0.763</b>	<b>0.803</b>		0.000			
Marquenas		0.000	0.000	0.000	0.000		0.000			
Lower Belt†		0.000	<b>0.086</b>	<b>0.164</b>	<b>0.173</b>		0.000	<b>0.805</b>	0.000	
Upper Belt†		0.000	<b>0.223</b>	<b>0.119</b>	<b>0.083</b>		0.003	0.025	0.001	0.001

Note: Only grains younger than ca. 1650 Ma were used in this comparison to assess the similarity of post-Mazatzal age populations. K-S statistic numbers indicate the probability (*P*) that two samples are derived from the same population. The higher the value, the more likely it is that the two age distributions were drawn from the same population. The *P* value must exceed 0.05 to be 95% confident that the two populations are not statistically different, and those values exceeding 0.05 are shown in bold and shaded in gray. Statistics were calculated using Microsoft Excel macros made available by G. Gehrels at the University of Arizona LaserChron Center (Gehrels, 2009).

\*Hess Canyon Group data from Doe et al. (2012) for the Yankee Joe Formation (YJ) and Blackjack Formation (BJ).

†Lower Belt composite includes data for the Piegan, Ravalli, and Lower Belt Groups and equivalents, and Upper Belt composite includes data from the Missoula and Lemhi Groups (Ross and Villeneuve, 2003; Link et al., 2007; Stewart et al., 2010).

in Ravalli Group and Piegan Group equivalents exposed in east-central Idaho. Deposition of the Lower Belt through Piegan Group occurred between  $1469 \pm 3$  Ma and  $1454 \pm 9$  Ma (Sears et al., 1998; Evans et al., 2000), indicating that the Belt-Purcell basin was active at the same time as basins in New Mexico and Arizona. Paleocurrent data and regional facies patterns from the lower part of the Belt succession indicate a western sediment source across the rifted and fault-segmented Laurentian margin (Winston, 1986; Ross et al., 1992; Ross and Villeneuve, 2003). Statistical comparisons of Mesoproterozoic detrital zircon ages from the lower part of the Belt succession with the Piedra Lumbre Formation and upper Hess Canyon Group samples indicate significant similarity (Table 2), raising the possibility that sediments from coeval portions of these basins were derived from similar sources.

Australia, Antarctica, and South America preserve significant records of magmatism and tectonic activity between ca. 1600 and 1500 Ma (Fig. 9A) and could have been proximal to Laurentia to the west or southwest (present coordinates) ca. 1490 Ma (Doe et al., 2012, and references therein). The North and South Australian cratons are the only known regions with a record of activity that spans 1629–1495 Ma (Betts and Giles, 2006; Cawood and Korsch, 2008), and the record of 1600–1580 Ma magmatism in associated regions of the Mawson continent in East Antarctica is growing (Peucat et al., 2002; Payne et al., 2009; Goodge et al., 2010). These occurrences closely match the range of potentially exotic detrital zircon ages contained in the Piedra Lumbre Formation and other contemporaneous successions in western North America (Fig. 9A), supporting plate reconstructions that place Australia and Antarctica west of North America ca. 1500 Ma (Karlstrom et al., 2001; Thorkelson et al., 2001;

Betts et al., 2002, 2008; Giles et al., 2004; Payne et al., 2009; Boger, 2011; Evans and Mitchell, 2011; Doe et al., 2012, 2013). We note that ca. 1600–1500 Ma sources were not present during deposition of the Marqueñas Formation sometime after ca. 1457 Ma or the Missoula Group starting at ca.  $1443 \pm 7$  Ma in the Belt-Purcell Basin (Evans et al., 2000). Instead, detrital zircons in these younger Mesoproterozoic successions are dominated by more local sources from southwestern Laurentia. This shift in provenance at ca. 1450 Ma may represent the final separation of the Australia–Mawson block and western Laurentia following ca. 1540–1500 Ma orogenesis in the North Australia craton and subsequent continental extension and erosion (Betts and Giles, 2006; Betts et al., 2007). We speculate, though, that it might also represent regional uplift in southern Laurentia related to the onset of contractional orogenesis.

### Unraveling Multiple Proterozoic Orogenies in Northern New Mexico

Resolution of the areal extent, intensity, and tectonic nature of the ca. 1650 Ma Mazatzal orogeny and ca. 1450–1400 Ma regional metamorphism and deformation is a longstanding research focus in the southwestern United States that has generated considerable debate (Grambling and Dallmeyer, 1993; Williams and Karlstrom, 1996; Karlstrom et al., 1997, 2004; Pedrick et al., 1998; Read et al., 1999; Williams et al., 1999; Shaw et al., 2005; Daniel and Pyle, 2006; J.V. Jones et al., 2010). Evidence of Paleoproterozoic greenschist-facies metamorphism and deformation is preserved locally where fabrics are crosscut by undeformed Paleoproterozoic granitic intrusions (Bauer and Williams, 1994) and where Paleoproterozoic plutons contain fabrics that are not observed in Mesoproterozoic plutons (Melis, 2001). Circa

1690 Ma  $^{40}\text{Ar}/^{39}\text{Ar}$  hornblende ages are also preserved in a greenschist-facies pluton in the Cimarron Mountains of northern New Mexico (Grambling and Dallmeyer, 1993). At a broader regional scale, evidence for the Mazatzal orogeny extends north to the Cheyenne Belt in southern Wyoming (Jones and Thrane, 2012) and northeast across the midcontinent to the Great Lakes region (Holm et al., 1998, 2007; Romano et al., 2000). Clearly, the Mazatzal orogeny was a regionally significant event.

However, our new detrital zircon data add to recent discoveries (Jones et al., 2011; Doe et al., 2012, 2013) that presumed Paleoproterozoic metamorphic rocks, in fact, have Mesoproterozoic protolith ages and require that deposition of the Trampas group and Marqueñas Formation occurred after ca. 1490 Ma. The ca. 1490–1460 Ma depositional ages, the similar deformation histories for both Paleoproterozoic and Mesoproterozoic units, an apparent lack of evidence for unconformities, and a predominance of metamorphic ages between ca. 1460 Ma and 1400 Ma all suggest that these supracrustal rocks experienced one amphibolite-facies regional metamorphic and deformational event during the Mesoproterozoic, contrary to previous models (Karlstrom et al., 1997, 2004; Karlstrom and Humphreys, 1998; Williams et al., 1999). Evidence for a clockwise *P-T* loop with peak pressures estimated at 0.3–0.6 GPa and kilometer-scale fold-and-thrust geometries in the 1490 Ma and younger metasedimentary units are consistent with significant crustal thickening at this time (Daniel and Pyle, 2006; Barnhart et al., 2012). Given the evidence for Mesoproterozoic deposition, deformation, regional metamorphism, and magmatism documented in the Picuris Mountains and across the reconstructed orogenic belt, we propose that this region records a significant orogenic event that we propose to call the Picuris orogeny. The initiation

of this orogeny is marked by deposition of the Pilar and Piedra Lumbre Formations and plutonic and volcanic activity in the southwestern United States as supported by the ca. 1480 Ma granitic plutons and meta-tuff layers. The end of this orogenic event is not well defined but may be associated with the end of amphibolite-facies metamorphism and waning regional plutonism between ca. 1400 and 1350 Ma.

We confidently extend the Picuris orogenic belt at least 40 km along strike to the east, where the Piedra Lumbre Formation is exposed in the Truchas Peaks and Rio Mora areas (Grambling and Codding, 1982), and west to the Tusas Mountains. We tentatively extend the orogen 400–500 km to the southwest (Fig. 1) on the basis of the strong similarity in detrital zircon age population between the Piedra Lumbre Formation and the deformed, weakly metamorphosed rocks of the Hess Canyon Group in Arizona (Doe *et al.*, 2012). Circa 1460 Ma deformation and metamorphism are also recognized up to 500 km south in the Burro Mountains of southern New Mexico (Amato *et al.*, 2011), and J.V. Jones *et al.* (2010) described ca. 1435–1365 Ma penetrative deformation and high-temperature metamorphism in the Wet Mountains to the north that they interpreted to represent pervasive lower-crustal flow. More localized reactivation of ductile shear zones ca. 1435–1380 Ma occurred up to 500 km to the north in Colorado (e.g., Selverstone *et al.*, 2000; Siddoway *et al.*, 2000; Shaw *et al.*, 2001; Jessup *et al.*, 2006; J.V. Jones *et al.*, 2010). Contrary to tectonic models that involve accretion of the Mazatzal Province along the southwest margin of Laurentia at ca. 1650 Ma, we speculate that final accretion may have occurred between ca. 1460 Ma and 1400 Ma, resulting in the Picuris orogeny.

## SPECULATIVE TECTONIC MODELS FOR MESOPROTEROZOIC DEPOSITION AND OROGENESIS

The hypothesis of a Mesoproterozoic (ca. 1450–1400 Ma) orogenic event for southern Laurentia is not new. Nyman *et al.* (1994) suggested that ca. 1400 Ma magmatism coincided with regional intraplate contraction arising from a convergent plate boundary on a distal southern margin of Laurentia. Regional evidence exists for contractional to strike-slip deformation within the thermal aureoles of ca. 1480–1380 Ma plutons and contemporaneous reactivation of northeast-striking crustal shear zones in the Rocky Mountains and southwestern United States, although the tectonic significance of these features is debated (Kirby *et al.*, 1995; Nyman and Karlstrom, 1997; Selverstone *et al.*, 2000; Karlstrom *et al.*, 2001; Shaw *et al.*, 2001;

Ferguson *et al.*, 2004; McCoy *et al.*, 2005; Jessup *et al.*, 2006; Amato *et al.*, 2011). This orogenic event was interpreted to be intracratonic following accretion of the Paleoproterozoic Mazatzal Province by ca. 1600 Ma (Karlstrom and Humphreys, 1998; Shaw and Karlstrom, 1999; Karlstrom *et al.*, 2004; Magnani *et al.*, 2004; Shaw *et al.*, 2005; Whitmeyer and Karlstrom, 2007; Jones *et al.*, 2009). Alternatively, we suggest that accretion of the Mazatzal Province might not have occurred until the Mesoproterozoic, and this orogenic event could reflect accretionary tectonic processes (Cawood and Buchan, 2007; Cawood *et al.*, 2009).

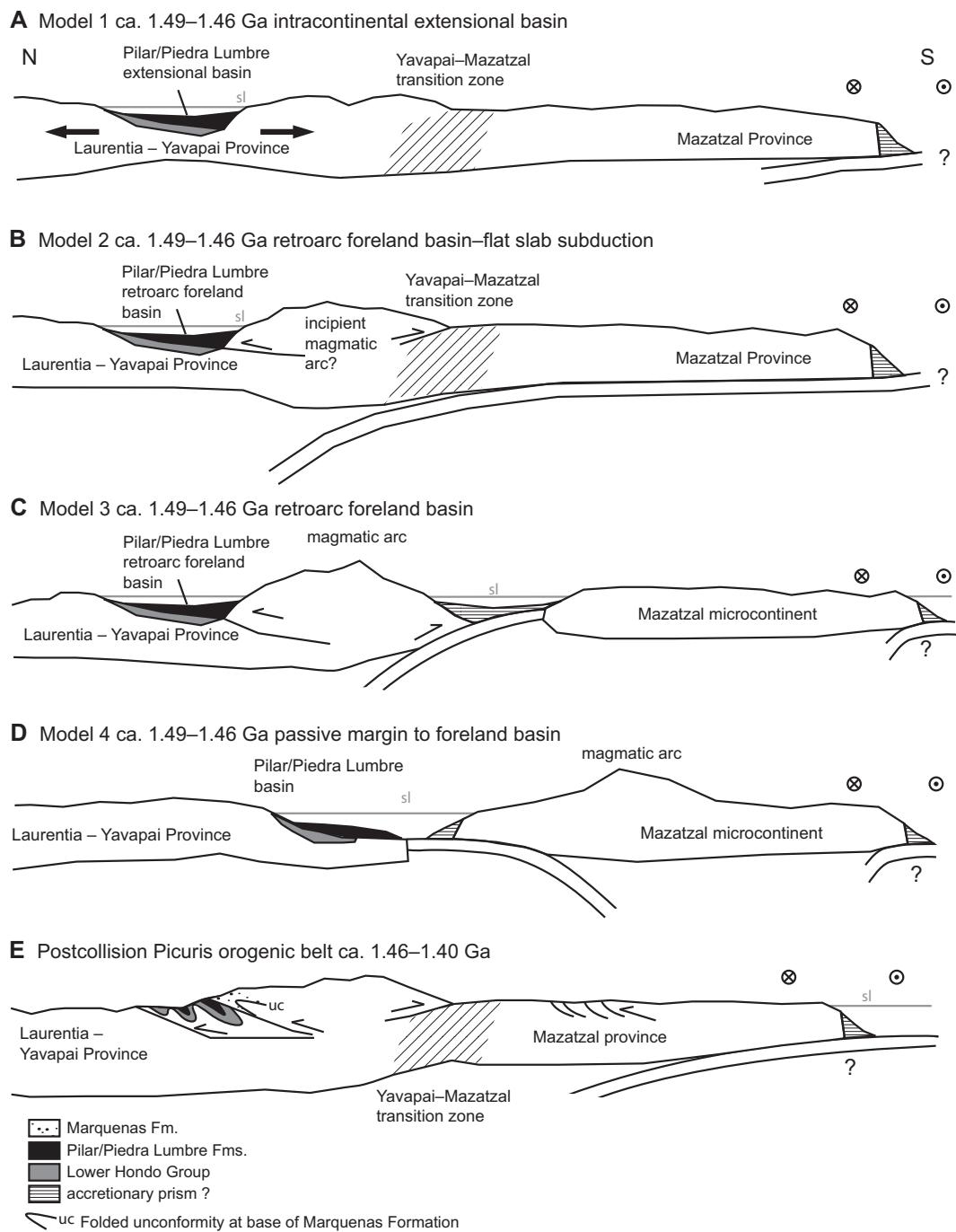
Figure 10 presents four models to attempt to explain Mesoproterozoic deposition of the Trampas group metasedimentary rocks and the subsequent structural and metamorphic evolution of rocks across New Mexico and southern Colorado. Models 1 and 2 show possible configurations that are based on accretion of the Mazatzal crustal province at ca. 1650–1600 Ma. Deposition of the Trampas group could possibly reflect intracratonic rifting (model 1; Fig. 10A) that produced a rift basin or aulacogen transecting the Yavapai-Mazatzal suture zone. Intracontinental rifting could have formed a basin network extending from New Mexico across Arizona to the continental margin, allowing for a mixture of detritus from the Yavapai and Mazatzal Provinces and exotic outboard sources. This extensional model for basin development is not favored given the minor occurrence of volcanic rocks interlayered with the clastic metasedimentary rocks, and the inferred sedimentary protoliths and depositional environments for the Pilar and Piedra Lumbre Formations.

Alternatively, model 2 (Fig. 10B) involves deposition in a retroarc foreland basin formed some 500 km inboard of the southern extent of the Mazatzal Province due to flat-slab subduction (Fig. 10B). As shown, crustal thickening occurred along the southern edge of the Yavapai Province, creating a barrier to detritus derived from the Mazatzal Province. In this scenario, the Pilar and Piedra Lumbre Formations would represent foredeep deposits, with the Marqueñas Formation representing proximal foredeep or wedge-top deposits thrust over the foredeep. Mazatzal-age detritus could have been supplied by ca. 1680–1650 Ma plutons that intruded the southern Yavapai Province and/or axial transport from Arizona and exotic source regions. Magmatism during basin formation would be limited during flat-slab subduction but could become pervasive as slab rollback or delamination occurred. Leaky transform faults and fracture systems within the flat slab could possibly allow for some magmatism to occur in the Mazatzal Province to the south.

The multipeak detrital age distribution of the Piedra Lumbre Formation and the inferred protoliths for the Pilar and Piedra Lumbre Formations are consistent with deposition along a continental margin in either a passive or convergent tectonic setting (Voice *et al.*, 2011). Figures 10C and 10D show two speculative tectonic models that invoke a continental convergent margin, where the southern edge of Laurentia transitions from a passive margin to a convergent margin beginning ca. 1490 Ma, due to the accretion of the Mazatzal Province. In model 3 (Fig. 10C), the Pilar and Piedra Lumbre Formations are interpreted to represent foredeep deposits in a retroarc foreland basin formed by flexural loading from a magmatic arc (DeCelles and Giles, 1996) that formed in response to north-dipping subduction along the southern edge of Laurentia. The coarsening-upward sequence observed in the Piedra Lumbre Formation is interpreted to reflect basin filling in an increasingly shallow marine setting followed by basin closure. The Marqueñas Formation would represent synorogenic wedge-top or proximal foredeep deposits that were thrust over the more distal foredeep deposits of the Pilar and Piedra Lumbre Formations. We envision an axial drainage pattern that transported sediment from southwest to northeast across central Arizona and into northern New Mexico (Fig. 9B). In this scenario, the developing magmatic arc would effectively block detritus from the ca. 1650 Ma Mazatzal Province. Detritus of this age could be derived from western sources or be more locally derived from 1680 to 1640 Ma plutons that intruded the Yavapai crustal province (Bauer and Williams, 1989; Bauer, 1993; Pedrick *et al.*, 1998). Following Soegaard and Eriksson (1985, 1986), the Ortega and Rinconada Formations are interpreted as being deposited in a passive-margin, shallow-marine and deltaic setting sometime between 1700 Ma and 1500 Ma. This model predicts the presence of a ca. 1500–1450 Ma accretionary prism south of the magmatic arc, unless the margin was sediment starved. Such a complex has not been recognized but could possibly be in the area of the Pedernal Hills and the northern Manzano Mountains.

In model 4 (Fig. 10D), the Pilar and Piedra Lumbre Formations are interpreted to represent relatively deep outer shelf to foredeep deposits in a passive margin or foreland basin formed by flexural loading from the colliding Mazatzal crustal province to the south. Southward subduction would create a magmatic arc that ultimately collided with the Yavapai Province margin during the Picuris orogeny; this scenario is analogous to the Taconic orogeny (Rowley and Kidd, 1981). The coarsening-upward sequence observed in the Piedra Lumbre Forma-

**Figure 10.** Speculative tectonic models to explain Mesoproterozoic deposition, regional metamorphism, and deformation in northern New Mexico. Models 1 and 2 assume Paleoproterozoic ca. 1650–1600 Ma accretion of the Mazatzal Province. Models 3 and 4 assume Mesoproterozoic accretion of the Mazatzal Province. (A) Model 1 represents basin formation and deposition of the Pilar and Piedra Lumbre Formations in an extensional intracontinental setting. (B) Model 2 shows Pilar and Piedra Lumbre Formations deposited within a retroarc foreland basin forming some 500 km inboard of the Mazatzal crustal province associated with flat-slab subduction. (C) In model 3, Pilar and Piedra Lumbre Formations are deposited within a retroarc foreland basin system. (D) In Model 4, Pilar and Piedra Lumbre Formations are deposited within a passive margin to foreland basin system. (E) Postcollision development of Mesoproterozoic fold-and-thrust belt across northern New Mexico, resulting in basin closure, deformation, and regional metamorphism of the Pilar, Piedra Lumbre, and Marqueñas Formations and older Paleoproterozoic rocks; sl—sea level.



tion would represent basin filling in an increasingly shallow marine setting followed by basin closure and thrust burial as evidenced by wedge-top or proximal foredeep deposition of the syntectonic Marqueñas conglomerate and quartzite. In this model, the basin could receive detritus from both the Mazatzal and Yavapai Provinces, in addition to exotic detritus by axial transport from the southwest (Fig. 9B). The Ortega and Rinconada Formations are interpreted as a shallow-marine-shelf passive-margin succession

(Soegaard and Eriksson, 1986) deposited sometime between 1700 Ma and ca. 1500 Ma.

One weakness of these models is that the development of a ca. 1490–1450 Ma magmatic arc is not well documented south of the Picuris Mountains. It is possible that remnants of the magmatic arc correspond to intermediate and granitic plutons exposed in the Santa Fe and Nacimiento Mountains and the Pecos complex (Robertson and Condé, 1989; Metcalf, 1990, 1995; Kellogg and Premon, 2005). Foliated

granodiorite in the Nacimiento Mountains gives a U-Pb zircon age of  $1453 \pm 9$  Ma (Kellogg and Premon, 2005), and the deformed Macho Creek granite in the Pecos complex is reported to have a U-Pb zircon age of 1480 Ma (no uncertainty reported; Robertson and Condé, 1989; Melis, 2001). There is no detailed geologic mapping or modern geochronology available in the northern Santa Fe Mountains to help evaluate this hypothesis. Circa 1400 Ma tonalite-diorite plutons in the central Santa Fe Mountains are

calc-alkaline to subalkaline, consistent with a subduction zone origin (Metcalf, 1990, 1995; Metcalf and Stropky, 2011).

In all of these models, ca. 1490–1460 Ma deposition is followed by basin closure and thrust burial, crustal thickening, deformation, magmatism, and regional metamorphism, producing the generalized structural geometries and relationships shown in Figure 10E. Paleo-proterozoic and Mesoproterozoic rocks in the southern Yavapai and Mazatzal crustal provinces were metamorphosed and penetratively deformed and show dominantly north-vergent structures (Fig. 10E). Deformation farther to the north in the Yavapai Province was strongly localized in the vicinity of coeval plutons or partitioned along reactivated shear zones (Selverstone *et al.*, 2000; Shaw *et al.*, 2001; McCoy *et al.*, 2005; Jessup *et al.*, 2006).

The main focus of this paper is to explain the timing and nature of Mesoproterozoic basin development and subsequent orogenesis, but any viable tectonic model must also account for regional ca. 1480–1360 Ma magmatism observed across the southern margin of Laurentia. The geochemical nature of this magmatic event is most compatible with a mantle plume or extensional event (Windley, 1993; Frost *et al.*, 1999; Anderson and Morrison, 2005). However, regional transpression could also, to some extent, account for both structural observations and the geochemistry of contemporaneous granites (e.g., Nyman *et al.*, 1994; Shaw *et al.*, 2001; J.V. Jones *et al.*, 2010). Models 1, 2, and 3 invoke north-dipping subduction with a potentially strong oblique component. In such a system, variation in subduction angle through time, episodes of slab rollback (Slagstad *et al.*, 2009), and possible mantle delamination could help to explain the observed regional patterns and characteristics of both magmatism and deformation in the overriding plate. Model 4 invokes an early south-dipping subduction zone that does not account very well for the early 1480–1450 Ma magmatism observed in the Yavapai crustal province to the north.

## CONCLUSIONS

New detrital zircon ages and a zircon crystallization age from a meta-tuff show that presumed Paleoproterozoic (ca. 1700 Ma) rocks of the Pilar and Piedra Lumbre Formations were instead deposited in the Mesoproterozoic ( $\leq$ ca. 1490 Ma). A single, robust tectonic model for Mesoproterozoic basin development, magmatism, deformation, and metamorphism remains elusive, but we favor a marine retroarc foreland basin model similar to models 2 or 3 (Fig. 10) and suggest that accretion of the Mazatzal Prov-

ince might have occurred more than 100 m.y. later than previously thought. The multipeak detrital age population for the Piedra Lumbre Formation is statistically indistinguishable from several detrital zircon age populations from coeval parts of the Hess Canyon Group in Arizona (Doe *et al.*, 2012) and the Belt Supergroup (Ross and Villeneuve, 2003; Link *et al.*, 2007), raising the possibility that the Arizona and New Mexico successions were part of a continuous and connected depositional system across southwestern Laurentia ca. 1490–1460 Ma. Circa 1500–1600 Ma detritus in the Piedra Lumbre Formation is inferred to have an exotic, non-Laurentian provenance, possibly from Australia or Antarctica. The youngest ca. 1450–1480 Ma zircon populations in the Piedra Lumbre and Marqueñas Formations are equivalent in age to plutons exposed in the region and suggest derivation from rapidly uplifted and eroded plutons or their eruptive equivalents and could reflect the westward extension of the Granite–Rhyolite province.

Our findings suggest that Proterozoic meta-sedimentary rocks in northern New Mexico experienced one major, regional episode of fold-and-thrust-style deformation and upper-amphibolite-facies metamorphism during the Mesoproterozoic. We do not dispute widespread evidence of ca. 1650–1600 Ma deformation throughout the region, but tectonic models that invoke ~200 m.y. of midcrustal residence for Proterozoic rocks of northern New Mexico following the Mazatzal orogeny are not supported by the findings of this study. The reconstructed Proterozoic fold-and-thrust belt documented across northern New Mexico (Karlstrom and Daniel, 1993; Daniel *et al.*, 1995; Cather *et al.*, 2006) consists of a series of kilometer-scale folds and ductile thrusts and regional metamorphic rocks that formed in response to a Mesoproterozoic (ca. 1490–1400 Ma) contractional to transpressional orogenic event that we propose to call the Picuris orogeny.

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